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FIELD DETERMINATION OF ICE JAM POROSITY

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ABSTRACT

Porosity is a variable frequently used in the analysis of river ice jams, such as in the direct calculation of jam volume, thickness, compressive strength, the coefficient of internal friction and ultimately in the prediction of water levels. Although a porosity of 0.4 is commonly assumed in the case of breakup jams, no field verification of this value has been made. This paper reports on the first calculation of ice jam porosity based on direct field measurements of the energy and mass balance of an ablating ice jam.

On April 29, 1983, an ice jam developed at the confluence of the Liard and Mackenzie Rivers and increased to a maximum length of approximately 22.8 km by May 4. After this time, discharge lowered, the toe of the jam remained stationary, and the jam decreased in length to only 6.3 km on May 9, before finally releasing into the Mackenzie River. Since flow velocities were too low to permit entrainment of ice beneath the jam, shrinkage of the accumulation was considered to be due to melt processes. Estimates of surface melt from atmospheric sources were made using known meteorologic and jam surface conditions. Notable, atmospheric heat inputs were less than 3% of those due to heat transferred by the incoming flow. Total jam volume was calculated using observed areal dimensions of the ice surface and an estimate of thickness obtained from application of the theory to floating, equilibrium jams. Ice jam porosity was then calculated by comparing the volume of ice melted within the jam to measured changes in the total (ice and water) jam volume.

RÉSUMÉ

La porosité est une variable souvent utilisée dans l'analyse des embâcles se formant sur les rivières et les fleuves pour calculer notamment leur volume, leur épaisseur, leur résistance à la compression et leur coefficient de frottement interne ainsi que pour prédire les niveaux d'eau. Bien que l'on évalue couramment la porosité d'un embâcle en fragmentation à 0,4, aucune vérification n'avait été effectuée sur place à cet effet. En mesurant sur place le bilan massique et énergétique d'un embâcle dont le volume diminue peu à peu, on est arrivé, pour la première fois, à en calculer la porosité de façon précise. Ce document décrit les résultats de cette recherche.

Le 29 avril 1983, un embâcle s'est formé au confluent de la rivière Liard et du fleuve Mackenzie. La glace s'est accumulée et le 4 mai, il a atteint sa longeur maximale, soit environ 22,8 km. Après cette date, le débit des cours d'eau a baissé, la partie de l'embâcle en aval est restée stationnaire et la longeur de ce dernier a diminué peu à peu (le 9 mai, il ne mesurait que 6,3 km). Plus tard, les glaces ont finalement été emportées par le courant du fleuve Mackenzie. Comme la vitesse d'écoulement était trop faible pour permettre l'effritement de la glace sous l'embâcle, on croit que cette baisse de volume a été causée par la chaleur. On a évalué le taux de fusion superficielle engendrée par les agents atmosphériques en tenant compte des facteurs météorologiques et des conditions à la surface de l'embâcle. Il est intéressant de noter qui est produite par le courant. Par ailleurs, on a calculé le volume total de l'embâcle en se basant sur la superficie observable de la couche de glace et l'épaisseur approximative de cette dernière, déterminée grâce à la mise en application de la théorie sur les embâcles flottants et en équilibre. On a ensuite calculé la porosité en comparant le volume de glace fondue aux variations observables du volume total (eau et glace) de l'embâcle.

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FIELD DETERMINATION OF ICE JAM POROSITY[1]

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ABSTRACT

Porosity is a variable frequently used in the analysis of river ice jams, such as in the direct calculation of jam volume, thickness, compressive strength, the coefficient of internal friction and ultimately in the prediction of water levels. Although a porosity of 0.4 is commonly assumed in the case of breakup jams, no field verification of this value has been made. This paper reports on the first calculation of ice jam porosity based on direct field measurements of the energy and mass balance of an ablating ice jam.

On April 29, 1983, an ice jam developed at the confluence of the Liard and Mackenzie Rivers and increased to a maximum length of approximately 22.8 km by May 4. After this time, discharge lowered, the toe of the jam remained stationary, and the jam decreased in length to only 6.3 km on May 9, before finally releasing into the Mackenzie River. Since flow velocities were too low to permit entrainment of ice beneath the jam, shrinkage of the accumulation was considered to be due to melt processes. Estimates of surface melt from atmospheric sources were made using known meteorologic and jam surface conditions. Notably, atmospheric heat inputs were less than 3% of those due to heat transferred by the incoming flow. Total jam volume was calculated using observed areal dimensions of the ice surface and an estimate of thickness obtained from application of the theory of floating, equilibrium jams. Ice jam porosity was then calculated by comparing the volume of ice melted within the jam to measured changes in the total (ice and water) jam volume.

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[1] NHRI Contribution No.88039

INTRODUCTION

The complex hydraulic and thermodynamic processes which control the formation, growth, decay and breakup of river ice produce a variety of ice forms and accumulations which vary widely in porosity (ratio of air and water pore volume to total volume). The porosity of ice jams is of particular interest because it is required for the direct calculation of, for example, jam volume, thickness, compressive strength, coefficient of internal friction and ultimately in the prediction of flood levels. Although a theoretical value of 0.4 is often assumed for the porosity of such accumulations (e.g. Beltaos 1983), no direct field verification of this value has been made. This paper reports on the direct calculation of jam porosity based on heat energy and mass balance computations associated with the thermal shrinkage of a breakup ice jam on the Liard River.

BACKGROUND THEORY

The porosity p_i of an ice jam can be expressed by:

$$[1] \qquad p_i = 1 - V_i/V_j$$

where V_i is the solid ice volume (m^3) and V_j is the total volume of the accumulation (m^3). The difference between these two volumes is the volume of voids filled with water (and air above the water line). Under melt conditions, changes in the total ice jam volume result in proportional decreases in the ice and void content. In a melt calculation, the total accumulation volume is given by:

$$[2] \qquad V_j = t_j L_j (B_j - 2B_s)$$

where V_j is the jam volume eliminated by melt processes (m^3), t_j is the jam thickness (m), L_j is the jam length (m), B_j is the jam width (m), and B_s is the width of ice remaining after melt (m), usually just shear walls which line the shores. The solid ice volume consumed by melt can be determined from:

$$[3] \qquad V_i = (\phi_w + \phi_a + \phi_g)/\rho_i \lambda_i$$

where ϕ_w is the hydrothermal heat transfer from water to ice ($W\ m^{-2}$), ϕ_a is the net atmospheric heat flux produced by sensible, latent and radiative heat transfers ($W\ m^{-2}$), ϕ_g is the ground heat flux to grounded ice ($W\ m^{-2}$), ρ_i is the density of ice ($kg\ m^{-3}$) and λ_i is the latent heat of fusion for ice ($J\ kg^{-1}$). In this study, ϕ_g is considered to be negligible and is therefcre not considered in further calculations.

The total flow of heat from water passing into an ice jam section can be expressed as:

$$[4] \qquad \phi_w = \rho_w C_w Q (T_i - T_w)$$

where ρ_w is the density of water (kg m^{-3}), C_w is the specific heat of water (J kg^{-1} $°C^{-1}$), Q is water discharge (m^3 s^{-1}), T_i is the accumulation ice temperature (°C) and T_w is the water temperature entering the jam reach (°C). In this case, it is assumed that all of ϕ_w is transferred to the ice accumulation. This is a reasonable assumption in view of the results of Marsh and Prowse (1987) which indicated that hydrothermal heat transfer to the base of rough jam surfaces is very rapid and produces rapid downstream decreases in T_w.

STUDY SITE AND HYDROMETEOROLOGIC CONDITIONS

a) Site Location

Ice jamming at the confluence of the Liard and Mackenzie Rivers at (61°53'N, 121°22'W) (Fig. 1) is an annual event and typical of jamming which occurs at the mouths of rivers which flow into lakes or into rivers that breakup later (Prowse 1986; Sherstone 1981). The morphological setting of the Liard River mouth, while clear of any sharp curves or bends, is highly conducive to jam formation. Within approximately 20 km of the confluence, the river slope decreases from approximately 0.00039 to only 0.00005, and over the last 10 km the river expands from a relatively constant open-water width of 700 m to almost 2800 m. The probability of jamming at the Liard mouth is further enhanced by the presence of intact ice on the larger Mackenzie River at the time of breakup on the Liard. In most years, flow from Great Slave Lake does not initiate breakup of the Mackenzie River upstream of the Liard River confluence until after the Liard jam has cleared into the downstream portion of the Mackenzie.

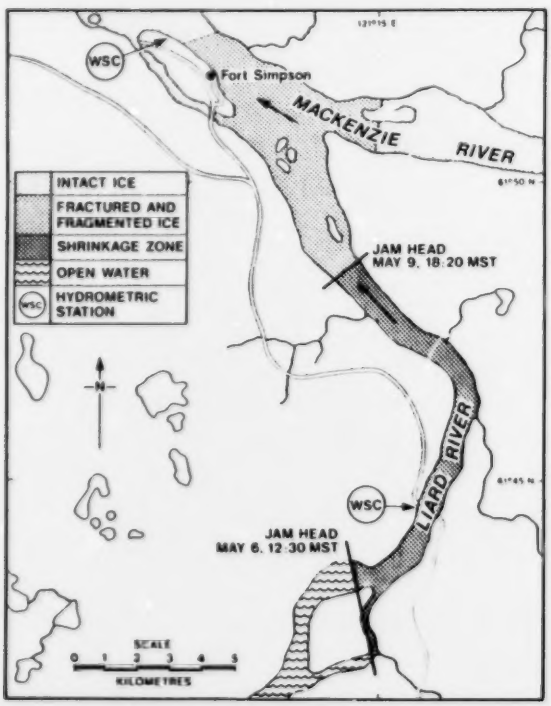

Fig. 1. Confluence of Liard and Mackenzie Rivers (1983).

b) Dimensions of Ice Jam and Channel

Duringthe1983 breakup of the Liard, an ice jam developed at the Liard mouth on April 29 and remained in place until May 10 before releasing and triggering breakup of the larger Mackenzie River (Prowse 1984). Large scale fracturing of the intact cover occcurred immediately downstream of the ice jam toe, but further downstream, near the Water Survey of Canada (WSC) hydrometric station on the Mackenzie River (Fig. 1), the cover remained largely intact. The ice jam reached a maximum length of 22.8 km on May 4 with the addition of upstream ice but then gradually began to reduce in length, down to only 6.3 km by May 9. Complete data for the calculation of V_i was only obtained from May 6 at 12:30 MST (Mountain Standard Time) to May 9 at 18:20 MST, during which the ice jam contracted 12.4 km. Over this period, the ice jam toe essentially remained stationary and shrinkage was only due to the downstream movement of the jam head. Average width of the jam was calculated to be 799 m based on five detailed cross-section profiles measured during open water conditions and water elevations obtained during the jam period. Using the theory of floating equilibrium jams, Prowse (1986) found the ice-jam thickness to be 5.63 m, a figure supported by similar heights of the shear walls remaining after the jam released.

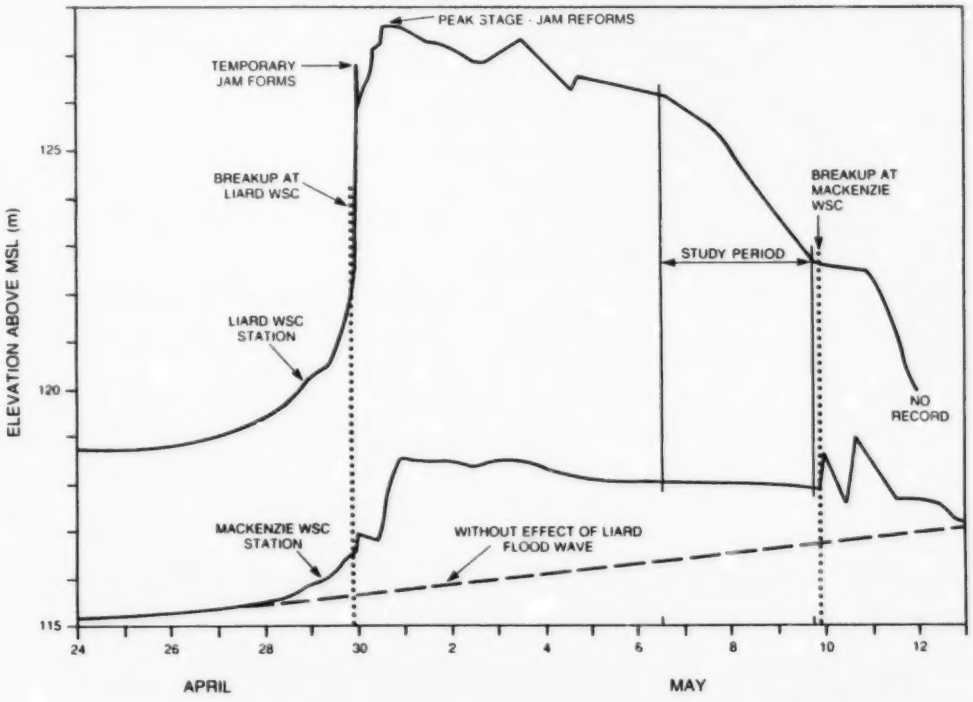

Fig. 2. Water levels during the 1983 ice breakup.

c) Flow conditions

Figs. 1 and 2 show the location of, and stage hydrographs for, the Liard and Mackenzie River hydrometric stations during the ice-jam event. Notably the Liard gauge is located in close proximity to the head of the Liard jam. During the 1983 breakup event, maximum stage was recorded at the Liard

on April 30 followed by a steady decline until late on May 9 when the Liard jam released into the Mackenzie River forming a second jam. Backwater from this second jam was largely responsible for the flattening of the stage record on the Liard ca. May 10. Over this period of declining stage, observations showed that ice was not being entrained beneath the ice jam, thus verifying that the jam shrinkage was due to melt and not the loss of ice to downstream sections.

Discharge from the Liard River contributes to the flow measured on the Mackenzie and is responsible for the almost simultaneous, although lower in magnitude, changes in stage recorded at the Mackenzie gauge. As noted by Prowse (1986), because only a slow rise in stage was observed to occur on the Mackenzie River upstream of the Liard confluence, the increase in stage at the Mackenzie gauge can be considered to be primarily due to the passage of the Liard flood wave.

Following the method outlined by Prowse (1986), discharge for the Liard station was calculated as being the difference in flow measured at the Mackenzie gauge with and without the Liard flood wave. The method is as follows; firstly, separation of the Mackenzie stage hydrograph was performed using an average, measured increase in pre-breakup stage of 0.11 m d^{-1}. The separation began on April 26, the day of the last discharge measurement on the Liard River. The two Mackenzie stage hydrographs were then used to calculate the Mackenzie flow with and without the Liard flood wave (Fig. 3).

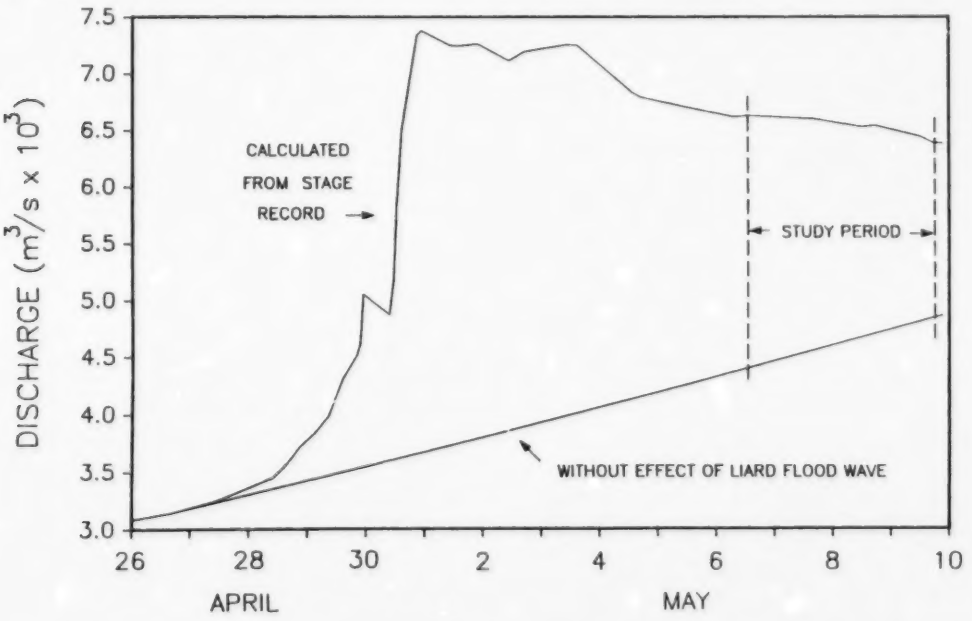

Fig. 3. Calculated discharges for the Mackenzie River (1983).

Both, however, include the baseline flow from the Liard of 825 m^3 s^{-1} measured on April 26. Discharge for the Mackenzie was calculated using:

$$[5] \qquad Q = B(4gS/f_o)^{0.5} (H - s_i t_i)$$

where g is the acceleration due to gravity (m s^{-2}), S is the water surface slope, f_o is the com-

posite friction factor for the ice-covered channel, H is the water depth (m), s_i is the specific gravity of ice and t_i is the thickness of the intact ice sheet (m). B was calculated from a stage-width relationship derived from cross-sectional surveys conducted over the zone of shrinkage; S was set equal to 0.00011, ice thickness surveys revealed t_i to be 1.0 m, and a value of 0.065 was employed for f_o based on the results of mid-April flow measurements.

Derived flow for the Liard River is shown in Fig. 4. The slow decline in the hydrograph after the peak on May 3 is believed to be due to the late breakup of major upstream tributaries such as the South Nahanni River and the Liard River above Nelson Forks (see Prowse 1984). The formula representing the decline in discharge during the study period of jam shrinkage is:

[6] $$Q = 2723 - 178t$$

where t is time (d) (R^2 = 0.99). Mean daily flow estimates made by the Water Survey of Canada are also shown in Fig 4. Although the two hydrographs attain similar maximums and reflect similar volumes of water, it is believed that the WSC trace underestimates the arrival time of the spring flood wave and overestimates discharge during the period of ice jam backwater.

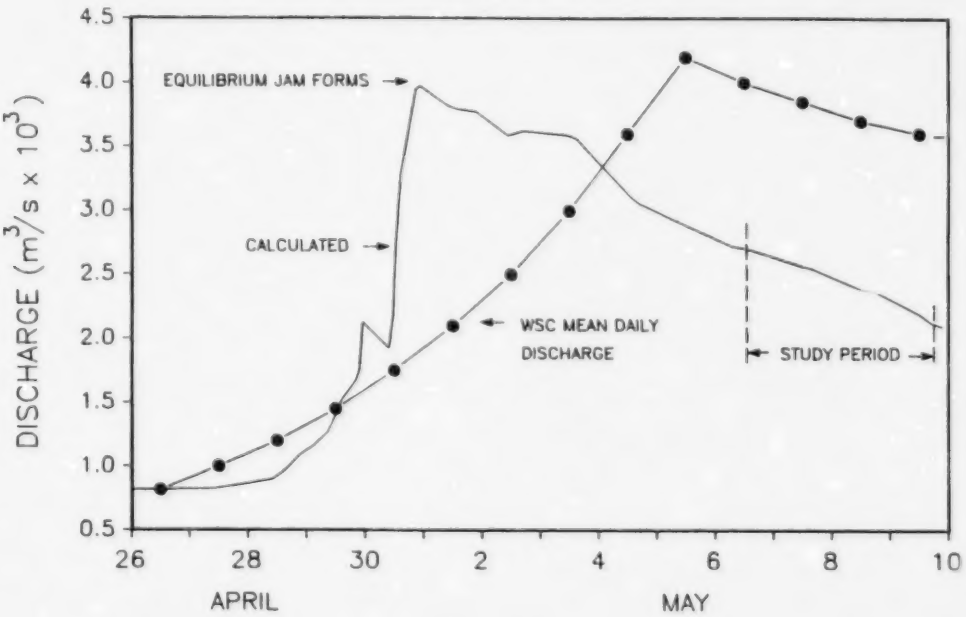

Fig. 4. Liard River discharge (1983).

d) Hydrothermal Heat Flow

With calculated Q, only T_w is required to estimate ϕ_w from [4] assuming that the ice accumulation temperature is at 0°C. This assumption is not unreasonable given the extensive fragmentation of ice and extensive mixing of the fragments within the river water prior to jam formation. Moreover, measurements conducted as part of a pre-breakup energy balance study, revealed that internal ice temperatures were only -0.2°C one day before breakup.

Beginning on May 6, water temperatures entering the jam were measured from a helicopter using an infrared radiation thermometer (sensitivity ±0.2°C at 20°C). Temperatures were taken at a number of locations within an approximate 5-km distance upstream of the ice jam head between 12:30 and 18:20 MST on May 6, 7 and 9. Although some diurnal fluctuation in T_w occurred, it is believed that temperatures obtained during this six-hour period approximated the daily mean within ±0.5°C. This is supported by the temperature data shown in Fig. 5 which was obtained under open-water conditions immediately following the ice-jam event. Mean daily water temperatures obtained for May 6, 7 and 9 are shown on Fig. 6 with those for a one-week period following breakup and a 0°C value for the day of breakup. The jam period data appear to fit an almost linear warming trend.

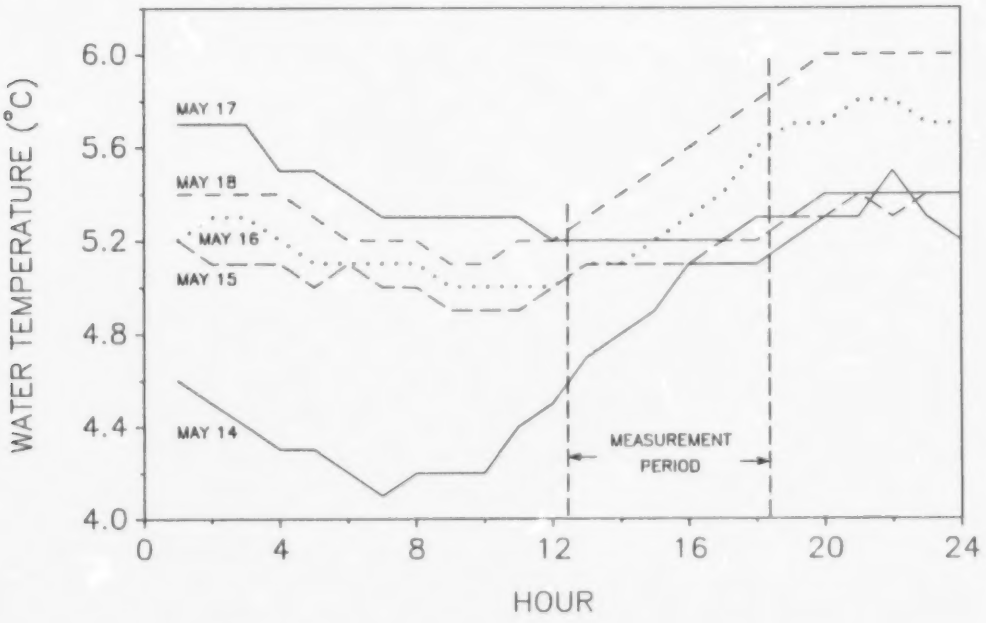

Fig. 5. Diurnal variations in water temperature (1983).

e) Atmospheric Heat Flow

Although no direct measurements of ϕ_a were obtained for the 1983 event, a reliable estimate can be made based on the results of a field study conducted at the same location in 1984 by Prowse and Marsh (1987). Some of the major conclusions from this study follow. During periods of intense melt, atmospheric heat flows to an intact ice cover usually outweigh those of an hydrothermal nature, particularly in cases where frazil ice is common. The magnitude of the ϕ_a even increases after the ice cover is fragmented by breakup processes. Prowse and Marsh (1987), calculated ϕ_a to be on average 145% greater under similar atmospheric conditions for fragmented as compared to intact ice. The dramatic increase is due to larger surface roughness, which increases the transfers of sensible and latent heat, and a lower albedo, which increases the amount of absorbed short-wave radiation. Prowse (1987), for example, notes the albedo of brash ice measured in an ice jam to average only 0.09. Under melt conditions, Prowse and Marsh (1987) found ϕ_a to range from approximately 4.2 to 16.7 MJ $m^{-2}d^{-1}$.

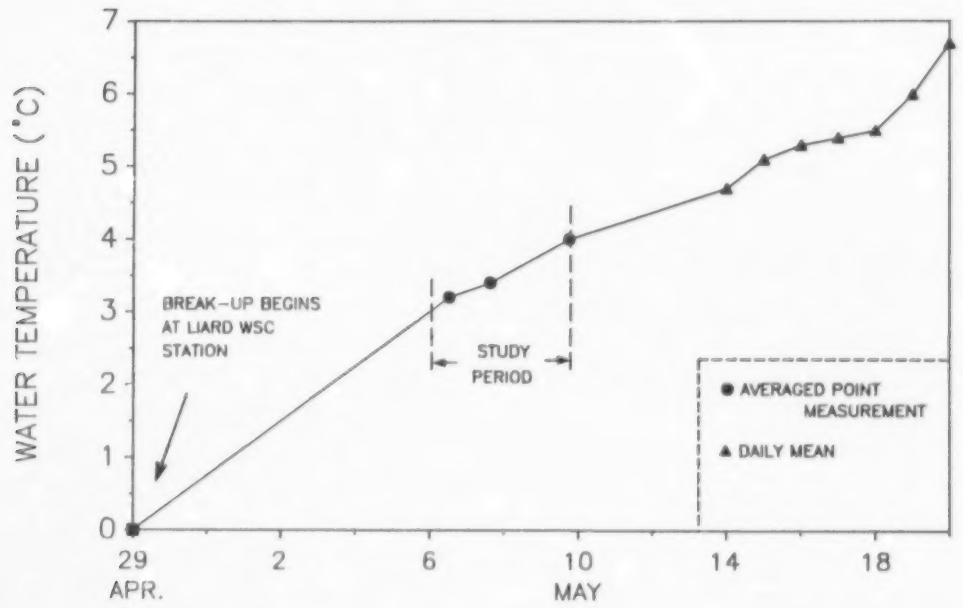

Fig. 6. Post-breakup water temperatures (1983).

An approximate value of ϕ_a can be obtained for 1983 according to an empirical equation established by Prowse et al. (1988) for intact ice conditions and inceased by the same 145% factor noted by Prowse and Marsh (1987) for brash ice:

[7] $$\phi_a = 4.36 + 0.67S_* + 1.07\bar{T}_a$$

where S_* is daily sunshine hours and $\bar{T}_a$ is the mean daily air temperature (°C)(R^2 = 0.63). Based on the meteorologic conditions prevailing over the study period, [7] gives a range of 5.7 to 11.5 MJ $m^{-2}d^{-1}$ and a weighted daily average for the study period of 9.1 MJ $m^{-2}d^{-1}$.

ANALYSIS AND DISCUSSION

Prior to estimating porosity, an assumption must be made about the jam configuration. As noted by Beltaos (1983), the streamwise profile of the bottom of an equilibrium jam includes three distinct sections: a downstream transition zone to intact ice, a middle equilibrium zone and an upstream transition zone. The shrinkage in the jam between May 6-9 is assumed to have affected the two latter zones only and that the downstream transition zone was contained within the lower 6.3 km of the jam. As earlier noted, Prowse (1986) calculated the equilibrium thickness to be 5.63 m and it is the equilibrium zone which is believed to occupy a majority of the 12.4-km shrinkage zone. Within the upstream transition zone, thickness increases from the head of the jam (upstream) approaching an asymptotic value equal to the equilibrium thickness. Unfortunately, thickness values are not available for this reach. However, such data are not required for the determination of porosity if the thickness profile within the upper zone can be assumed to be the same at the end of the shrinkage

period as at the beginning. This assumption is reasonable considering the similar shape of the thickness profile expected to result from hydrothermal melt. As suggested by the results of Marsh and Prowse (1987), rapid transfers of heat to the bottom of jams produces a rapid temperature decay curve which should also be reflected in the thickness profile. In view of the above, the equilibrium jam thickness was employed in [2] to calculate $V_j = 5.21 \times 10^7$ m^3.

Calculation of ϕ_w in [4] was based on a calculated mean water temperature of 3.6°C and integration of [6] over the shrinkage period. This resulted in $\phi_w = 1.02 \times 10^{10}$ MJ which when combined in [3] with $\phi_a = 2.77 \times 10^8$ MJ for the entire melt surface gives $V_i = 3.40 \times 10^7$ m^3. Note that ϕ_a is less than 3% of ϕ_w. When the heat flow figures are combined in [1] with the above value of V_j, p_i = 0.35. As mentioned earlier, the major uncertainty in the calculations lies with the water temperature measurements. Allowing the mean temperature to vary by ±0.5°C results in a p_i range of 0.26 to 0.44, with the latter value representing cooler water as perhaps indicated by the temperature trend in Fig. 5. Although the difference between the calculated value of p_i and its commonly assumed value of 0.4 is relatively small, it would seem prudent to re-evaluate the various other ice jam parameters which rely on ice jam porosity in their calculation and to examine the sensitivity and potential range in each based on the values of p_i derived in this analysis.

CONCLUSION

Final calculation of p_i has required the combination of a number of hydraulic and meteorological calculations which rely on detailed sets of field information which are difficult to collect, particularly during periods of ice breakup. The data, however,are believed to be the best, and probably the only, available and provide a good first approximation of ice jam porosity. Notably, and encouragingly, the field-derived porosity of 0.35 is very close to the laboratory value of 0.4, a figure which has been commonly assumed in a variety of ice engineering formulae without field verification.

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DISCUSSION SHEET

Discussed by : Steve Daly

on Paper by : Terry Prowse

Paper No. : B34: Field Determination of Ice Jam Porosity

QUESTION

Measurements made by a impulse radar on a brash ice jam on the St. Clair River indicates that the porosity above and belov the water line in the jam may not be the same. The porosity appears to be greater above the water line. One possible mechanism to cause this would be "sifting" of small pieces in the ice above the water line, downward to the water line. Do you have any comments on the possible non-uniform porosity in the jams you study?

DISCUSSION SHEET

Discussed by : Steve Daly

on Paper by : Terry Prowse

Paper No. : B34: Field Determination of Ice Jam Porosity

REPLY

No direct measurements or calculations were made of vertical or streamwise variations in porosity within the Liard brash ice accumulations. However, the existence of lower porosity above the water line does seem possible for the Liard case, and in fact for most jam situations. This is especially true where the "sifting" of ice particles is aided by surface melt, which leads to the fragmentation of ice floes above the water line and subsequent development of a less porous accumulation of smaller pieces near the water surface.

Vertical non-uniformity of porosity beneath the water line has also been measured for the case of ice distributed by the passage of ships (e.g. Sandkvist, 1981). The sifting of floes, as the fragmented ice attains a buoyancy equilibrium after the ship passage, is a likely explanation for the porosity variations. Similarly, it is possible for the case of "narrow-channel" river ice jams that variations in porosity develop because the ice floes are sifted as they flow into the ice jam reach. The sifting of particles leading to significant porosity variations beneath the water line is less likely for the case of "wide-channel" jams, especially those produced by dynamic or mechanical break-ups. Rapid downstream shoving and thickening leaves little opportunity for the sifting of particles. If large variations do exist within "wide-channel" jams, some of the assumptions employed in this theory will have to be re-evaluated.

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Additional notes by author:

Note that in the energy-mass balance calculations, the symbols ϕ_a and ϕ_w are referred to as heat fluxes (W m^{-2}; in Eq. 3 and 4), as total quantities of heat transferred per day (MJ m^{-2} d^{-1}; Eq. 7) and as total heat transferred to the entire accumulation over the entire study period (MJ; calculation of V_i and p_i).

To avoid confusion the following equations should appear as:

[3] $V_i = A_j \, \Delta t(\phi_a + \phi_w + \phi_g)/\rho_i \, \lambda_i$

where A_j is the area of melt (m^2), Δt is the duration of the melt period (s), ϕ_a is the net atmospheric heat flux produced by sensible, latent and radiative heat transfers (W m^{-2}), ϕ_w is the hydrothermal heat flux from water to ice (W m^{-2}), ϕ_g is the ground heat flux to grounded ice (W m^{-2}), ρ_i is the density of ice (kg m^{-3}) and λ_i is the latent heat of fusion for ice (J kg^{-1}).

[4] $\phi_w = \rho_w \, C_w \, Q \, (T_i - T_w)/ \, A_j$

where ρ_w is the density of water (kg m^{-3}), C_w is the specific heat of water (J kg^{-1} .°C^{-1}), Q is discharge (m^3 s^{-1}), T_i is the accumulation ice temperature (°C) and T_w is the water temperature entering the jam reach (°C).

[7] $\phi_a = 4.36 + 0.67S_\star + 1.07T_a$

where ϕ_a is the atmospheric heat flux (MJ m^{-2} d^{-1}), $S_\star$ is the daily sunshine hours and T_a is the mean daily air temperature (°C).

Note also that B_s in Eq. 2 refers to the width of a single shear wall and assumes equal widths of ice on the left and right banks.